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Switch-off of a major enhanced ice flow unit in East Antarctica

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[1] The East Antarctic Ice Sheet (EAIS) is the largest reservoir of ice on the planet by an order of magnitude. Compared with the West Antarctic Ice Sheet (WAIS), it is traditionally considered to be relatively stable, with only minor adjustments to its configuration over glacial-interglacial cycles. Here, we present the results of a radio-echo sounding survey from Coats Land, East Antarctica, which suggests that parts of the EAIS outlet drainage system may have changed significantly since the Last Glacial Maximum. We have identified an enhanced flow unit from buckled internal layering and smooth bed morphology that is no longer active. We believe this feature to have shut down at some point since the Last Glacial Maximum, ~20 ka BP. **Citation:** Rippin, D. M., M. J. Siegert, J. L. Bamber, D. G. Vaughan, and H. F. J. Corr (2006), Switch-off of a major enhanced ice flow unit in East Antarctica, *Geophys. Res. Lett.*, 33, L15501, doi:10.1029/2006GL026648.

1. Introduction

[2] It is now widely accepted that instead of an abrupt transition between streaming and inland flow in ice sheets [Alley and Whillans, 1991], the Antarctic ice sheet exhibits an extended zone of transition between these two end members, in which flow occurs by some combination of internal deformation and basal motion [Joughin *et al.*, 1999]. These so-called ‘tributaries’ have been shown to be widespread throughout both East and West Antarctica [Bamber *et al.*, 2000], and their presence has important implications for understanding the flow and stability of ice sheets.

[3] The Slessor Glacier complex in Coats Land, East Antarctica (Figure 1) is an increasingly well studied area [e.g., Rippin *et al.*, 2003a, 2004; Bamber *et al.*, 2006; Rippin *et al.*, 2006]. Previous work has shown that enhanced flow tributaries here lie in deep, well-defined, basal troughs that are separated from each other by areas of higher ground. The large-scale location of the tributaries is, therefore, well constrained [Rippin *et al.*, 2003a]. Combined synthetic aperture radar interferometry (InSAR) and balance velocities indicate that maximum velocities in the tributaries are of the order of ~60 m a⁻¹ [Bamber *et al.*, 2006]. Calculations have indicated, however, that the contribution of basal motion to tributary flow can vary between tributaries:

flow in two of the three tributaries of Slessor Glacier can largely be explained by ice deformation (with basal motion, within the range of uncertainties, playing a minor role), while flow in the third (under a similar glaciological setting), could only be explained by significant basal motion [Rippin *et al.*, 2003a].

[4] Radio-echo sounding (RES) data-sets from many regions of Antarctica have revealed that internal layers are common and traceable continuously over >100 km (e.g., Robin and Millar, 1982; Siegert, 1999; Hodgkins *et al.*, 2000). Internal layers occur as a result of electromagnetic reflections at boundaries of dielectric contrast, and are attributed to changes in ice density, variations in acidity as a consequence of atmospheric aerosols from volcanic eruptions, or to changes in crystal orientation fabrics [Siegert and Fujita, 2001]. In some instances, layers have been observed to become disrupted and buckled. Continuous layers are generally found in slow-moving inland regions where there has been little ice motion since deposition, while layers appear distorted and buckled over very rough beds, or in areas of faster flow such as ice streams and their tributaries [Robin and Millar, 1982; Jacobel *et al.*, 1993; Bell *et al.*, 1998; Rippin *et al.*, 2003b; Siegert *et al.*, 2003].

[5] Here, we present the results of an airborne RES campaign carried out in the upper reaches of Slessor Glacier. In this paper, we investigate the englacial stratigraphy and identify differences in the nature of this layering. We interpret these results with respect to information regarding ice dynamics and bed morphology, in an effort to shed light on the flow configuration of this part of the EAIS since the Last Glacial Maximum (LGM).

2. Methodology

[6] An airborne RES campaign was carried out in the upper reaches of Slessor Glacier during the austral summer of 2001/2002. A field camp and fuel depot were located at 078° 58.60’S, 007° 24.97’W, from where five airborne RES sorties were flown, over a total distance of ~5000 km, arranged in a grid measuring 200 × 280 km (Figure 1).

[7] The British Antarctic Survey (BAS) airborne radar is a coherent system, mounted on a de Havilland DHC-6 Twin Otter aircraft. It transmits a modulated pulse at a centre frequency of 150 MHz and with a peak power of 1200 W. A constant aircraft speed of 60 m s⁻¹ was maintained, giving an along-track sampling resolution of ~60–75 m. The across-track spacing of the radar-lines was 40 km. Surface elevation data were provided by the aircraft’s on-board avionics radar system, which recorded terrain-clearance more accurately than the RES system itself. Sorties were flown between pre-defined waypoints (Figure 1), and GPS coordinates were recorded using an on-board Ashtech GPS receiver, with reference to a second receiver at the field

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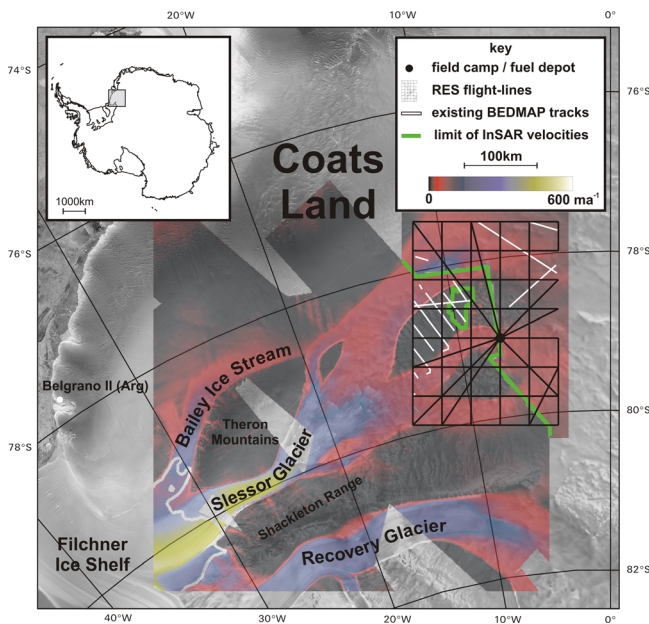


Figure 1. The RADARSAT Antarctic Mapping Project (RAMP) mosaic of Antarctica [Jezek and RAMP Production Team, 2002], in Coats Land. Slessor Glacier is clearly visible, as well as the adjacent Bailey Ice Stream and Recovery Glacier. Flight-lines and the field camp/fuel depot are marked. Where they are available, Interferometric Synthetic Aperture Radar (InSAR) velocities are displayed as a color overlay (I. Joughin, personal communication, 2002). Where InSAR velocities are not available, balance velocities are instead shown. The boundary between InSAR and balance velocities is shown as a green line (InSAR down-stream of the line, balance velocities up-stream). The location of the base map in Antarctica is shown in the inset figure.

camp. Initial processing of the RES data was carried out using the seismic processing package ProMAX v2003.0. The location of the surface and bed were identified using a semi-automatic picking process. GPS data were processed using the Trimble package GPSurvey. Processing procedures are described in more detail given by Rippin *et al.* [2003a, 2004, 2006].

[8] Once processing had been completed, the RES data were classified in terms of the type of internal layering visible as follows: i) well preserved continuous internal layers which are either flat, or deformed in line with underlying subglacial topography; ii) layers which are buckled or disrupted in such a way that does not reflect subglacial topography; and iii) a lack of internal layers [Rippin *et al.*, 2003b; Siegert *et al.*, 2003]. Although it was straightforward to identify regions that lacked layers, we could not be certain that this absence occurred due to glaciological phenomena (i.e., destruction of layering due to extreme buckling processes perhaps). It is possible that, in some cases, layers were not visible due to a reduction in returned radar power as a consequence of increased ice thickness or changing aircraft altitude. As a result of this uncertainty, we did not interpret areas which lacked layers [Rippin *et al.*, 2003b; Siegert *et al.*, 2003]. These classified data were then compared with bed morphology and ice flow

indicators to assess the connection between ice dynamics and englacial structures.

3. Results

[9] Figure 2 shows the bed topography in the Slessor Glacier region [Rippin *et al.*, 2003a]. The distinct troughs associated with the three tributaries – Slessor Tributary North (STN), Slessor Tributary Central (STC) and Slessor Tributary South (STS) – are clearly visible (and marked). Furthermore, there is another large, deep area toward the south-east of the study area. This trough lies to the east of the heads of STC and STS and is termed the *deep south-eastern trough* (DSET) [Rippin *et al.*, 2006].

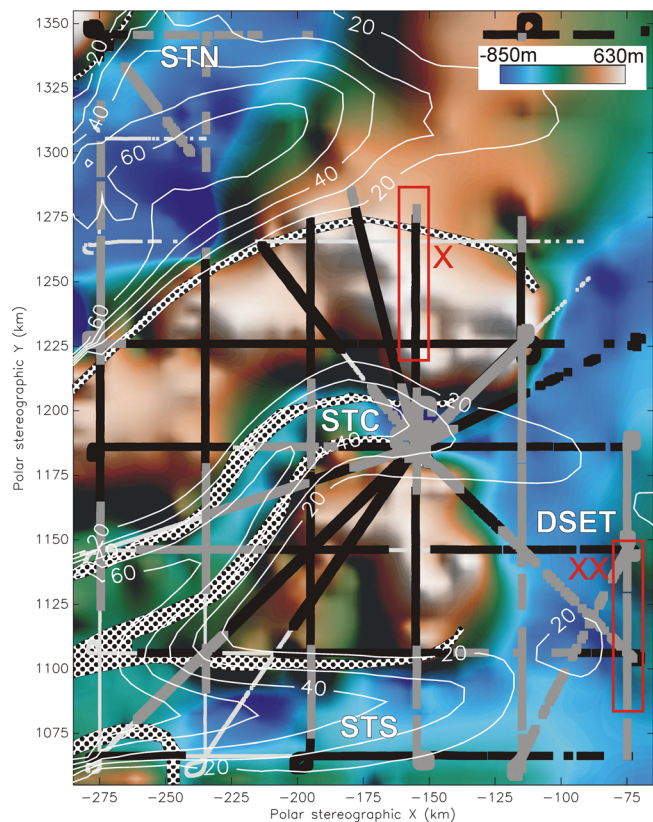


Figure 2. Bed topography in the Slessor region of East Antarctica [after Rippin *et al.*, 2003a], shown in color. The three troughs containing the tributaries of Slessor Glacier are clearly visible, and marked as STN (Slessor Tributary North), STC (Slessor Tributary Central) and STS (Slessor Tributary South). DSET is the deep south-eastern trough. The flightlines are marked, to aid location with reference to Figure 1. Thick black lines represent areas of continuous internal layering along flight-lines, while thick grey lines represent areas where layers are buckled. Thin grey lines indicate where bed returns were achieved, but where no internal layers are visible. Contours represent a combination of InSAR velocities and balance velocities (m a^{-1}). Shear margins, delineating the limits of enhanced flow, derived from the RAMP mosaic [Jezek and RAMP Production Team, 2002], are marked by black dots. 'X' and a thin rectangular box marks the location referred to in Figure 3, while 'XX' and a thin rectangular box marks the location referred to in Figure 4.

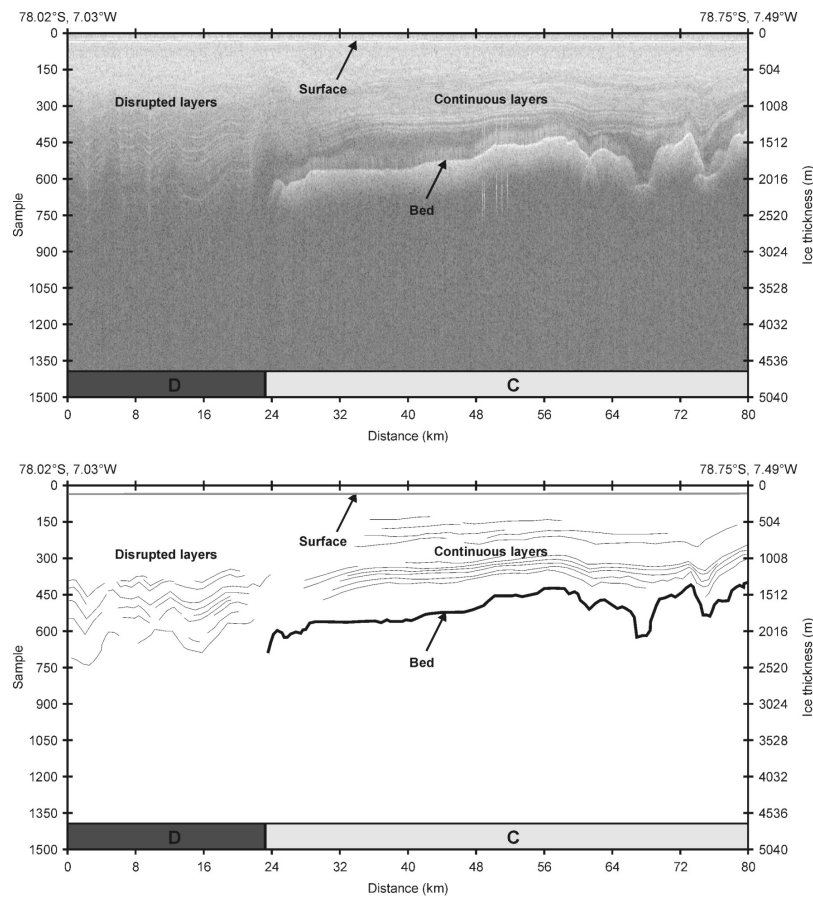


Figure 3. Example of a boundary between continuous internal layers and disrupted layers. (top) Data in grey-scale format, while (bottom) traced layers are marked. In both figures, the surface and bed are indicated, as are areas of continuous and disrupted layers. The extent of continuous layers ('C') and disrupted layers ('D') are indicated by shaded bars at the bottom of the figure. The location of the figure is marked by a red rectangular box and an 'X' on Figure 2.

[10] It should be noted that in our original analyses, the northernmost trough (STN) was incorrectly described as a tributary of the nearby Bailey Ice Stream [e.g., *Rippin et al.*, 2003a, 2004]. However, more recent inspection of Interferometric Synthetic Aperture Radar (InSAR) velocity data clearly indicates this to be a tributary of Slessor Glacier. (e.g., *Bamber et al.*, 2006; *Rippin et al.*, 2006].

[11] Figure 2 also shows the location of different types of layering along the flightlines. All inter-tributary areas are dominated by continuous internal layers, while all the enhanced flow tributaries are dominated by disrupted internal layers, and the boundaries between continuous and disrupted layers at the margins of enhanced flow tributaries are sharp. These boundaries coincide with shear-margins which are visible on the RADARSAT Antarctic Mapping Project (RAMP) mosaic (Figures 2 and Figure 3) [*Jezek and RAMP Production Team*, 2002]. At the head of STC the dense array of flightlines indicates a significant area of buckled internal layering (Figure 2).

[12] Elsewhere, areas of continuous layering are evident in parts of the very upper reaches of STS, while DSET is dominated by disrupted internal layers (Figures 2 and 4), although there are also prominent regions here where layers are continuous (Figure 2). Finally, there is a lack of layers in some of the margins between slow and fast flow, and in the

lower reaches of both STC and STS. There is also a significant lack of layering in much of STN (Figure 2).

4. Discussions

[13] All layer-types found in the slow-moving inter-tributary areas, and in the Slessor enhanced flow tributaries can be understood with respect to present day ice dynamics. Disrupted layers in the Slessor tributaries imply velocities here are sufficient to set up stresses of the required magnitude to disrupt internal layers [*Siebert et al.*, 2003]. Such extensive buckled internal layering may be caused by flow convergence/divergence in tributaries of fast flow [*Ng and Conway*, 2004] and/or elevated longitudinal stresses within areas of increased velocities, and increased shear stresses at the margins between areas of fast and slow flow [*Jacobel et al.*, 1993; *Siebert et al.*, 2003]. Disruption of internal layers has been widely reported in ice streams, but reports from enhanced flow tributaries are far fewer [*Rippin et al.*, 2003b; *Siebert et al.*, 2003].

[14] It is significant that there are disrupted internal layers in all tributaries, since previous work has indicated that the mechanism responsible for flow in STN is markedly different to that responsible in STC and STS. Approximately 70% of the movement of STN is explained by basal motion, while virtually all motion in STC and STS can be explained

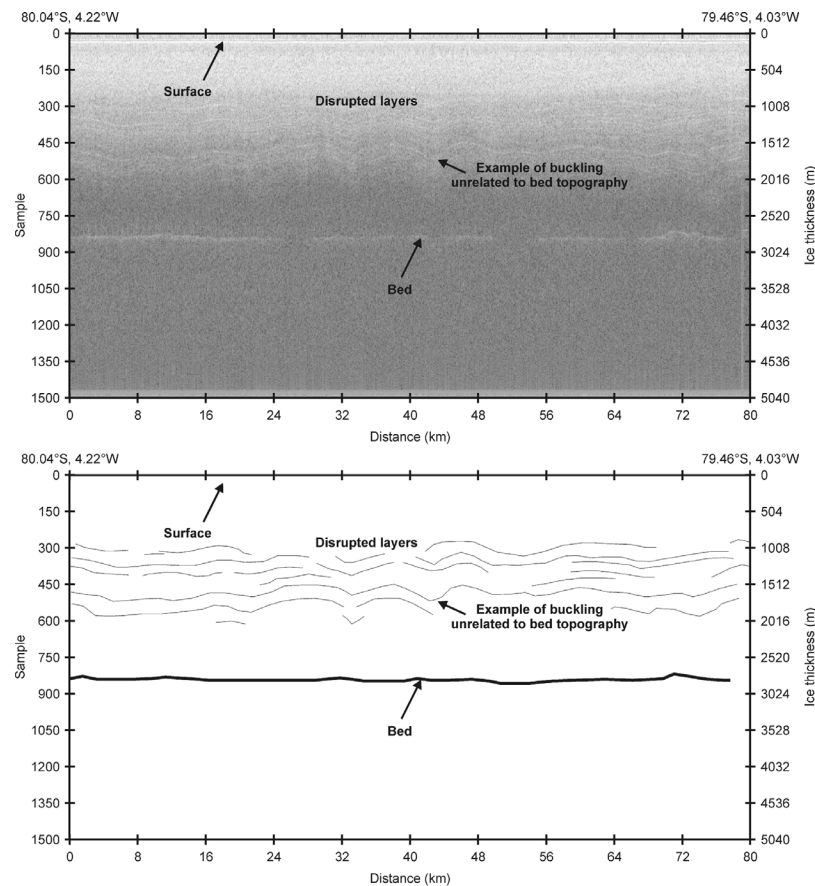


Figure 4. Example of distorted internal layers in the DSET. (top) Data in grey-scale format, while (bottom panel) traced layers are marked. In both figures, the surface and bed are indicated, as well as the distorted layers. Note how smooth the bed is. The location of the figure is marked by a red rectangular box and an ‘XX’ on Figure 2.

by ice deformation [Rippin *et al.*, 2003a]. Siegert *et al.* [2003] suggest that it is the difference in the dominant flow mechanism that explains why layers are disrupted in the tributaries of Ice Stream D (where basal motion is important), but are continuous and well preserved in the Bentley Subglacial Trench (where ice deformation dominates), inland of the Siple Coast region of West Antarctica. The data presented here from East Antarctica indicates, however, that velocities in enhanced flow tributaries are sufficiently high to lead to layer disruption, even where all this flow may be explained by deformation alone.

[15] In contrast to the Slessor tributaries, the DSET is an area of thick ice (~ 2800 m), where estimated flow velocities (derived from combined InSAR and balance velocities) are generally low (largely ~ 10 m a^{-1}) and there is a lack of basal motion [Rippin *et al.*, 2003a; Bamber *et al.*, 2006]. The region is, however, dominated by disrupted internal layers (Figure 2 and 4). Such disruption is clearly associated, in other parts of Antarctica, with enhanced ice flow [cf. Rippin *et al.*, 2003b]. Consequently, it appears the process that caused the layers to buckle has been replaced with slow flow [Ng and Conway, 2004]. This area of East Antarctica has therefore undergone significant dynamic change, and the disrupted layers observed there cannot be understood with respect to present day dynamics. We propose that the existence of disrupted layers in currently

slow moving ice indicates that this trough was occupied by fast-flowing ice in the past.

[16] There is further evidence to support this proposition. Rippin *et al.* [2004, 2006] have shown that the bed in DSET is particularly smooth, which is taken to indicate that rapid flow occurred in the past, and either led to the presence of a smooth bed (through processes of subglacial erosion), or that the existence of a smooth bed helped facilitate past rapid flow here. Figure 4 shows a section from the DSET, in which the bed is seen to be very smooth, yet there is significant disruption of the layering above it, unrelated to the bed topography.

[17] In order to determine the date when the buckling process occurred, we employed the approach of Dansgaard and Johnsen [1969] to date particular layers. This approach uses a flow model to date layers, using a constant vertical strain rate down to a distance of 400 m above the bed (beneath an ice thickness of 3000 m), and then allows the strain rate to decrease linearly to the bed below [Dansgaard and Johnsen, 1969; cf. Paterson, 1994]. Layering of any sort is difficult to see in the very upper and lower parts of the radar data, and so there is uncertainty as to whether the upper and lower boundaries between buckled and unbuckled layers have been precisely identified. Nevertheless, we attempt to date the lowest (oldest) and highest (youngest) buckled layers, to determine the earliest date the buckling could have begun, and latest it could have ended.

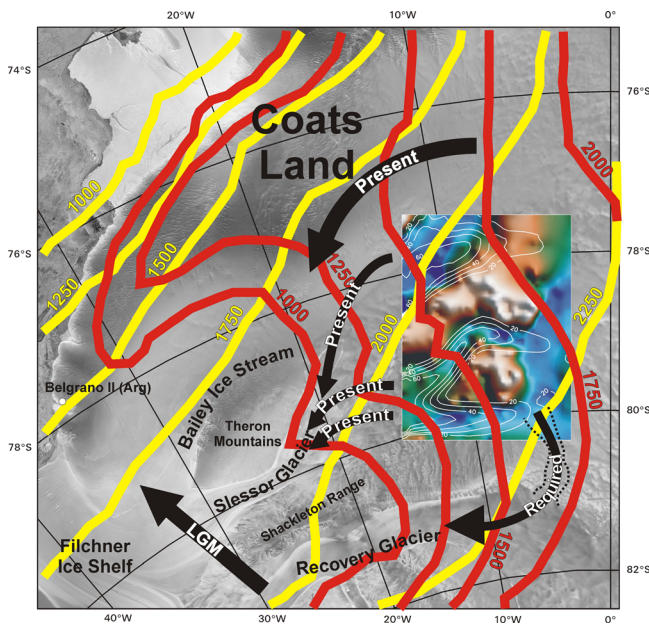


Figure 5. Reconstructions of surface elevation (m) in relation to present sea level at the LGM (thick yellow contours) and at present (thick red contours), derived from Huybrechts [2002]. Approximate flow direction at LGM and at present, based on surface elevation, is shown by black arrows, and labeled ‘LGM’ and ‘present’ respectively. The black arrow labeled ‘required’ indicates the flow direction we believe is necessary to explain the existence of the observed buckled layers and the smooth bed in the DSET. The dotted black lines surrounding this required flow direction indicate the location of faintly visible surface features linking the DSET with Recovery Glacier. Contours overlie part of the RAMP mosaic [Jezek and RAMP Production Team, 2002], and in the Slessor tributary region, bed topography is also displayed in color, and combined InSAR and balance velocities in thin white contours (see Figure 2).

Using this method, we determine that the buckling process could have begun up to ~ 60 ka BP (the date of the deepest buckled layers), and continued until at least ~ 18 ka BP (the date of the uppermost buckled layers). Thus the process is not related to the last interglacial period (~ 130 ka BP).

[18] Allowing for uncertainties in our dating of the buckling process, from these results, we suggest that buckling may have been at its peak during the Last Glacial Maximum (LGM), ~ 20 ka BP [Watanabe *et al.*, 2003] when the ice sheet was thicker [Huybrechts, 2002], and as a consequence, velocities were greater. Indeed, Huybrechts [2002] has shown that there may have been significant changes in flow orientation in this part of East Antarctic Ice Sheet (EAIS) at the LGM (Figure 5) [cf. Huybrechts, 2002]. His work demonstrates that at the LGM, the Filchner-Ronne region was more expansive, which appears to have dragged Coats Land with it [Huybrechts, 2002]. As a consequence, surface slopes and flow directions at the LGM were markedly different to those in the present day (Figure 5). This reconstruction means that ice in Coats Land would have been thicker, and would have fed from a different direction. Crucially however, despite a significant change in

flow direction, Huybrechts’ [2002] proposed change in ice-sheet configuration at the LGM is not able to explain the apparent flow direction required to cause the observed buckling and smooth bed in the DSET. If our calculations are accurate, then it draws into question the applicability of Huybrechts’ [2002] reconstruction at the LGM in this region. Our work requires that, at the LGM, ice sheet configuration was such that ice flowed southward through the DSET, in order to enable faster flow here. We propose that the DSET may in fact have fed into Recovery Glacier, to the south of Slessor Glacier (Figure 5). This is supported by faintly visible features in the RAMP mosaic of Antarctica [Jezek and RAMP Production Team, 2002], linking the DSET with the upper reaches of Recovery Glacier (marked on Figure 5). We propose that these features may be relict shear zones delineating the margins of past fast flow, and thus further supporting our interpretation of the DSET as having once carried faster-flowing ice. However, if these are indeed relict shear zones, then their current visibility implies that faster flow is likely to have occurred here more recently than ~ 20 ka BP, otherwise they would have been buried. In addition, because flow velocities here are of the order of 10 m a^{-1} , over 20 ka, disrupted layering would be transported ~ 200 km. The DSET is ~ 150 – 200 km in length, and so if the layers were indeed disrupted as a result of palaeo-fast flow in the same location, then 20 ka BP would seem to be a likely absolute maximum age for the shut-down. We are not able to provide any more accurate idea of the date of faster-flow shutdown, but the evidence available indicates this is likely to have occurred in the last few ka, more recently than the LGM.

[19] We cannot be certain whether merely increasing ice thickness would have been enough to raise velocities sufficiently to explain buckled layering in the DSET. It is thus possible that basal sliding may also be a requirement. However, it is important to note that we observed buckled layers in all the Slessor tributaries, even where basal motion is not significant [cf. Rippin *et al.*, 2003a]. This observation is in contrast to previous work, as explained previously [Siebert *et al.*, 2003]. Thinning of ice during the Holocene (and possible basal freeze-on) may have been responsible for the slowing of ice in the DSET since then.

5. Conclusions

[20] In this paper, we have mapped internal layering in Coats Land, East Antarctica, and shown that over much of the region, continuous layers exist in slow-moving intertributary regions, while fast-flowing tributaries are dominated by disrupted layers. Additionally, other work analyzing basal roughness also revealed that in tributaries where basal motion dominates, the bed is smooth, while where basal motion is not significant, the bed is rough [Rippin *et al.*, 2004, 2006]. These observations indicate that most of the layering observed can thus be related to modern flow conditions. However, there is one region that is anomalous. The DSET currently exhibits slow flow, but possesses a smooth bed and disrupted layers. We propose that this trough once contained fast-flowing ice, as a consequence of large-scale changes in the form of the ice sheet and in resultant ice flow direction at the LGM, which would have influenced the location of major drainage outlets. However,

the changes in ice sheet configuration at the LGM that are required to explain this change in flow direction are greater than those proposed by Huybrechts [2002]. Over the last ~20 ka since the LGM, the ice sheet configuration has been modified to its current state and fast-flow in this region has shut down.

[21] Reorganization of ice stream configuration has been reported before, from the Siple Coast region of West Antarctica [e.g., Christoffersen and Tulaczyk, 2003; Conway et al., 2002; Catania et al., 2005]. Relict flow features associated with past fast flow have also been identified, and such changes have been interpreted as a consequence of changes in surface topography in response to the effects of internal dynamics and climate change [Conway et al., 2002]. Although very significant changes in flow direction were reported [Conway et al., 2002], these changes are often explained by changes in the location of the boundary between frozen and thawed bed, in relation to the minimal topographic controls on ice stream location, and thus their ability to migrate laterally [Catania et al., 2003]. The work presented here from EAIS, however, differs because of the deep subglacial troughs that seem likely to provide significant topographic control to the location of enhanced flow features, preventing this lateral migration. However, despite this topographic influence, flow direction in this part of East Antarctica seems to have changed markedly since the LGM, and we believe that this may explain the apparent relict fast flow in the DSET.

[22] The work presented here is the first recorded observation of such temporal changes in ice stream dynamics in East Antarctica, and is of significance because it suggests that that ice flow in East Antarctica is subject to considerable variation in flow regime over glacial-interglacial cycles [cf. Huybrechts, 2002] and that at the LGM (and probably more recently), enhanced flow extended further inland than at present. Such changes to the EAIS are of significance in understanding the history of the ice sheet and its reaction to long-term climate change, particularly as the EAIS is traditionally considered to be relatively stable.

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